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1 **Origin and provenance of 2 Ma – 2 Ga zircons ejected by phreatomagmatic**
2 **eruptions of Pliocene basalts in southern Slovakia**

3
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13

14 **Abstract:** U-Pb ages of zircons recovered from Pliocene pyroclastic deposits in northern part
15 of the Cenozoic intra-Carpathian back-arc basin (Pannonian Basin) span the interval from
16 Pliocene (2.2 Ma) to Paleoproterozoic (Orosirian-Rhyacian, 1850-2115 Ma). The scattered U-
17 Pb ages reflect eruption ages of the host basaltic volcanic centres, two episodes of post-Eocene
18 magmatic crustal growth, and the possible tectonic affiliation, provenance and age of the
19 subjacent basement or the sedimentary basin detritus sampled by the basaltic magma.

20 The youngest zircons define maximum ages of phreatomagmatic eruptions during the Late
21 Miocene-Pliocene extension. These zircons are distinguished from older zircons by Zr/Hf (40-
22 90) and Th/U ratios (0.5-4.5) as well as super-chondritic $\varepsilon_{\text{Hf}}(t)$ values ranging from +7 to +14,
23 indicating mantle-derived parental magmas. The locally increased Th/U ratios (up to 8)
24 accompanied by Zr/Hf >60 are diagnostic of evolved phonolite parental melt. Hence, the
25 youngest zircons can be interpreted as antecrysts, originating from evolved melts cogenetic

26 with the host alkali basalts. In contrast, older zircons represent xenocrysts scavenged by the
27 uprising basalt from surrounding rocks.

28 Subordinate Eocene-Early Oligocene (29-38 Ma) sub-group of zircon xenocrysts is
29 coincidental with the magmatism and volcanism along the Periadriatic lineament and the
30 middle-Hungarian zone. The Early Miocene (18 Ma) cluster is coeval with the deposition of
31 the Bükk Mountains felsic ignimbrite correlated with the onset of the back-arc extension that
32 triggered Miocene sedimentation within the Pannonian Basin. The Eocene-Early Oligocene
33 zircons have been likely scavenged from pyroclastic and ash-fall deposits of the Palaeogene
34 retroarc basin subjacent to the Miocene basin infilling. Sub-chondritic $\epsilon\text{Hf}(t)$ values between -
35 2.5 and -8 in the Eocene-Early Miocene zircons attest their crystallization from subduction-
36 related felsic-to-intermediate melts containing large amounts of recycled crustal material.

37 Palaeozoic-Proterozoic zircons create a heterogeneous population with variable trace element
38 abundances and $\epsilon\text{Hf}(t)$ values. The determined age clusters are reminiscent of some basement
39 units cropping out recently in Central Western Carpathians. Zircon Hf isotope data indicate
40 recycling of up to 3.4 Ga old mafic crust and also the presence of 2 Ga old juvenile mafic crust.
41 These units had either underlain the northern part of the Pannonian Basin during Pliocene, or
42 had been exposed during the deposition of Miocene clastic sediments. The absence of
43 Mesoproterozoic, Grenvillian zircons (0.9-1.8 Ga) in the pre-Cenozoic population of zircon
44 xenocrysts is provisionally interpreted as indicating the evolution of the zircon source area
45 within the west-African Craton.

46

47 *Keywords:* basalt, zircon, U-Pb-Hf isotopes, Pannonian Basin, Western Carpathians

48

49 **1. Introduction**

50

51 Zircon megacrysts, several mm to cm in diameter, often occur in Zr-undersaturated mafic
52 magmas related to continental and oceanic intra-plate settings (Hoskin and Schaltegger 2003).
53 The megacrystic zircons provide invaluable information addressing eruption ages of host
54 basalts (Siebel et al. 2009, Hurai et al. 2013), and oxygen isotope composition, temperature and
55 Nd-Hf systematics of subcontinental mantle (Valley et al. 1998, Paquette and Mergoïl-Daniel
56 2009, Siebel et al. 2009, Schaltegger et al. 2015). There is a consensus that the megacrystic
57 zircons in mafic igneous rocks are in fact antecrysts crystallized from mantle-derived alkalic
58 melts differentiated at various depths as felsic-to-intermediate intrusive bodies and transported
59 to surface in younger basalt portions (e.g. Aspen et al. 1990; Guo et al. 1996; Upton et al. 2009;
60 Paquette and Mergoïl-Daniel 2009; Yu et al. 2010; Sutherland et al. 2016; Schmitt et al. 2017).
61 The term antecryst refers to zircon crystals crystallized from an earlier pulse of magma
62 incorporated in a later pulse (Miller et al. 2007). In contrast, old zircons with different
63 geochemical characteristics incorporated from the surrounding host rocks or those unrelated to
64 the magmatic system are named xenocrysts (Johnson 1989).

65 Megacrystic zircons have also been detected in products of phreato-magmatic eruptions of
66 the south-Slovakian Volcanic Field (SSVF) extending over northernmost promontories of the
67 intra-Carpathian back-arc basin in the Lučenec Basin and the Cerova Highlands. Hurai et al.
68 (2012, 2013) showed that the zircon megacrysts typologically and chemically resemble those
69 present in syenite xenoliths from the same pyroclastic deposits (Huraiová et al. 1996; Hurai et
70 al. 1998). Homogeneous U-Pb ages of zircons from maar sediments showed uninterrupted
71 growth in sub-volcanic reservoirs within ± 100 ka. Only one of the studied zircon megacrysts
72 showed heterogeneous zoning created by a 2.55 ± 0.14 Ma old core and 1.91 ± 0.07 rim, which
73 reflected a $\sim 650 \pm 200$ ka long crystallization in a magmatic reservoir. Older inherited cores have
74 not been observed (Hurai et al. 2010, 2012). Apatite and zircon crystals recovered from the
75 same pyroclastic deposits returned ages almost equivalent (within uncertainty limits) closure
76 ages of the U-Pb system at magmatic temperature (>900 °C) and the low-temperature (~ 240 -

77 60 °C) post-eruptive cooling recorded by the (U-Th)/He system (Hurai et al. 2013), thus
78 attesting the antecrystic character of the studied zircon megacrysts.

79 U-Pb isotopes have been hitherto determined in zircons from a total of five maars in the
80 SSVF (Hurai et al. 2010, 2012, 2013). This study has been extended to additional maar-diatreme
81 structures with the aim to better characterize not only megacrystic, but also sub-megacrystic
82 zircons smaller than 200 µm. We provide new U-Pb and Lu-Hf isotope data from a total of 44
83 zircons recovered from maar-diatreme structures from SE part of the SSVF. Compared to
84 previous results, the expanded dataset paints more complex picture of scattered ages, involving
85 zircon antecrysts and older zircons xenocrysts containing also inherited cores. This paper is
86 aimed to set the new U-Pb and Lu-Hf isotope data into geological context.

87 In other parts of the Carpathian back-arc basin, U-Pb, Lu-Hf, (U-Th)/He isotope or fission
88 track zircon data are only limited to Middle Eocene-Early Oligocene tuffs in western Pannonian
89 Basin (Danišík et al. 2015) and Early-to-Middle Miocene ignimbrites in Bükk Mts. in northern
90 Pannonian Basin (Lukács et al. 2015, 2018). U-Pb ages of zircons in a metasedimentary
91 granulite xenolith isolated from the basaltic tuff of Balaton Highlands in the central part of the
92 Carpathian back-arc basin (Downes et al. 2015) recorded a complex thermal history correlated
93 with late Mesozoic (90-100 Ma) and Palaeogene (40-65 Ma) metamorphism in the adjacent
94 orogenic collisional belts, the increased heat flow at 15-16 Ma coincidental with the Miocene-
95 Pliocene overheating and the final eruption of the host basalt at 3.7 Ma in Pliocene.

96

97 **2. Geological setting**

98

99 The south-Slovakian Volcanic Field (SSVF) is located within the Alpine-Carpathian-
100 Pannonian (ALCAPA) mega-unit (Fig. 1a) composed of the Alpine-Carpathian orogenic belt
101 and the superimposed intra-Carpathian back-arc basin (Pannonian Basin). The ALCAPA mega-
102 unit is separated from the neighbouring Tisza-Dacia mega-unit to the south by the middle-

103 Hungarian zone, which represents the continuation of the Periadriatic lineament. Both mega-
104 units were juxtaposed during Late Oligocene-Early Miocene lateral extrusion of the ALCAPA
105 from Alpine collision zone (Kázmér and Kovács 1985, Csontos 1995). During Paleogene, the
106 area experienced subduction accompanied by sedimentary basin formation in foreland, forearc
107 (Kázmér et al. 2003) and retroarc (Tari et al. 1993) of the Alpine-Carpathian orogenic belt.
108 Superimposed back-arc extension during Early Miocene controlled by the subduction of
109 oceanic lithosphere underlying former flysch basins, and the ongoing subduction in Eastern
110 Carpathians (Nemcok et al. 1998; Konečný et al. 2002) triggered the back-arc sedimentation of
111 the Pannonian Basin starting at ~19 Ma (Royden et al. 1983), accompanied by intense volcanic
112 activity.

113 Subduction with roll-back, slab-break-off, delamination, strike-slip tectonics and block
114 rotation in the ALCAPA region has been accompanied by several episodes of syn- and post-
115 collisional magmatic activity (Nemcok et al. 1998; Konečný et al. 2002; Seghedi et al. 2004;
116 Harangi et al. 2006; Pécskay et al. 2006; Harangi and Lenkey 2007; Kovács et al. 2007; Lexa
117 et al. 2010): 1) Eocene-Early Miocene volcanism activated by dextral strike-slip faulting along
118 the Periadriatic lineament, subdivided into subordinate Eocene (60-37 Ma) and more extensive
119 Late Eocene-Early Oligocene (37-25 Ma) and Late Oligocene-Early Miocene (25-17.5 Ma)
120 pulses; 2) Early Miocene (21-11 Ma) calc-alkaline rhyolite-dacite volcanism coeval with early
121 extensional phases; 3) areal- and linear-type middle-Miocene calc-alkaline andesite to bimodal
122 andesite-rhyolite volcanism associated with the partial melting of mantle wedge above
123 subducting slab (Szabó et al. 1992) and/or to late-stage back-arc extension (Lexa and Konečný
124 1974); and 4) disseminated Upper Miocene-Pleistocene Na-alkaline basaltic and basanitic
125 volcanism (11.5-0.1 Ma) coincidental with the partial melting of mantle triggered by the
126 asthenosphere bulging and crustal thinning (Embey-Isztin and Dobosi 1995). The thinned crust
127 and lithospheric mantle in the study region is recently underlain in the depth of ~70 km by an

128 asthenospheric bulge (Tašárová et al. 2009; Bielik et al. 2010) uplifted to a depth of ~22 km in
129 the central part of the Pannonian Basin (Horváth 1993).

130 The SSVF covers an area of about 500 km² along the Hungary-Slovakia state boundary
131 (Fig. 1b). The SSVF is the classical region, where xenoliths of mantle peridotites (Hovorka and
132 Fejdi 1980; Szabó and Taylor 1994; Huraiová and Konečný 1994; Konečný et al. 1995a;
133 Downes and Vaselli 1995, Liptai et al. 2017) and granulites (Kovács and Szabó 2005) have
134 been studied to decipher the composition and properties of lower crust and mantle, as well as
135 processes associated with the mantle metasomatism and melting. Evidences for basaltic magma
136 underplating, fractionation within mantle and crustal reservoirs, and generation of anorogenic
137 magmas have been obtained in this area from the study of clinopyroxene and amphibole-bearing
138 mafic cumulates (e.g. Huraiová et al. 1996; Zajacz and Szabó 2003; Kovács et al. 2004; Hurai
139 et al. 2007; Zajacz et al. 2007) as well as syenites and A₁-type intra-plate granites (Huraiová et
140 al. 1996, 2005, 2017a, b, 2019; Hurai et al. 1998, 2010).

141 The SSVF basalts span compositions from basanite through basaltic trachyandesite to
142 phonotephrite. Trace element distribution patterns and radiogenic isotopes in the basalts
143 (Embey-Isztin and Dobosi 1995) combined with those in mantle xenoliths (Downes and Vaselli
144 1995; Liptai et al. 2017) reflect a 7-25 % partial melting of spinel-bearing mantle source with
145 basalt compositions locally affected by the enriched (metasomatized) mantle.

146 The SSVF basalts erupted within the interval from ~7 Ma to ~200 ka during several
147 successive phases (Konečný and Lexa 2007). The initial Late Miocene phase was identified in
148 NW part of the SSVF (Balogh et al. 1981). Superimposed Pliocene-Quaternary basalts erupted
149 in SE part of the SSVF during 2nd to 6th consecutive phases (Konečný et al. 1995b). Phreatic
150 and phreato-magmatic eruptions triggered and accelerated by the interaction of basalt with
151 aquifers gave rise to maar-diatreme structures formed by Surtseyan, Strombolian and Hawaiian
152 eruptions (Lexa et al. 2010). Subsequent vertical tectonic movements along NW–SE- and NE–

153 SW-trending faults exposed diatremes – feeder conduits of maars destructed by erosion
154 ([Konečný et al. 1999](#)).

155 U-Pb zircon dating ([Hurai et al. 2010, 2012](#)) complemented with low-temperature (U-
156 Th)/He geochronometry applied to zircon and apatite ([Hurai et al. 2013](#)) proved an Upper
157 Miocene (~5.5 Ma) eruption age of maars near Pinciná and Fiľakovo–Castle Hill, and Pliocene
158 eruption ages of maars near Hodejov (2.9 Ma), Hajnáčka (3.1 Ma) and Gemerské Dechtáre (3.2
159 Ma). Zircons extracted from a granite xenolith in the lava flow near Čamovce village ([Huraiová
160 et al. 2017b](#)) returned an U-Pb age of 5.18 ± 0.02 Ma similar to the 5.2 ± 0.9 Ma whole-rock
161 K/Ar age of the host basalt ([Konečný 2007](#)), thus attesting the capability of the U-Pb zircon
162 geochronometer to constrain the basalt eruption ages.

163

164 **3. Samples**

165

166 Zircons have been recovered from two diatremes (Šurice, Tachty) and one maar (Fiľakovo–
167 Red Hill). An isolated cliff in the southern margin of Šurice village named Owl’s Castle
168 ($48^{\circ}13'34.04''\text{N}$, $19^{\circ}54'44.94''\text{E}$, [Figs. 2a, b](#)) is an exposed feeding channel of deeply eroded
169 maar containing a block of subsided maar sediments in the upper part. The early infilling of the
170 diatreme consists of palagonite tuffs and breccias with larger blocks and fragments of Early
171 Miocene sandstones. The subsided block of stratified maar deposits is composed of tuffaceous
172 siltstones and sandstones grading in the uppermost part into palagonite/tuff breccias with
173 agglutinated scoria fragments and bombs. The sedimentary succession reflects the transition
174 from early Surtseyan wet-surge to late Strombolian eruptions. The diatreme infilling as well as
175 the subsided maar sediments are cut by basalt dykes, representing feeder channels of Hawaiian
176 eruptions (lava fountains) and possibly of an eroded crater lava lake. The K/Ar whole rock
177 dating of the basalt dyke returned an age of 4.0 ± 0.3 Ma ([Konečný et al. 2003](#)), although the
178 diatreme itself has been affiliated with the 3rd volcanic phase lasting from 2.45 to 2.75 Ma

179 (Konečný 2007). The volcanoclastic material from the Šurice diatreme exhibits regular
180 distribution of grain size fractions and significant volume of coarser grain-size fractions,
181 indicative of sorting in a maar lake.

182 The diatreme in western part of Tachty village (48°9'21.07"N, 19°56'42.40"E, Figs. 2c, d)
183 is filled with well-stratified, partly welded basaltic tuff containing sandstone fragments and
184 blocks of the early Miocene Fil'akovo Formation (Burdigalian) subjacent to the tuff cone. The
185 pyroclastic succession is dominated by lapilli tuffs with glassy lapilli embedded in siliciclastic
186 matrix. Variable amounts of sideromelane and tachylite reflect a complex magma
187 fragmentation history, recycling in the feeder conduit and interactions with water (Németh et
188 al. 2011). The sharp sub-vertical contact of the diatreme wall with the caustically modified
189 Burdigalian sandstone is preserved along the eastern side of the outcrop. The Tachty diatreme
190 has also been affiliated with 3rd volcanic phase (Konečný 2007). Another small outcrop exposed
191 in southern part of the Tachty village, 200 m SW from the previous sampling site, contains
192 pyroclastic material mixed with larger sandstone fragments with green glauconite grains
193 diagnostic of an Early Miocene sandstone/arcose. The sample studied is an unsorted mixture of
194 volcanoclastic and sedimentary material ejected during a phreato-magmatic eruption from the
195 proximal volcanic centre. Two zircon grains were recovered and studied from this site.

196 The Red Hill maar (48°16'43.27"N, 19°49'24.76"E) is situated in northern part of the
197 Fil'akovo town. Similar to the proximal maar located in the centre of Fil'akovo (Castle Hill), it
198 is built of a relict tuff ring of coarse-grained, unsorted wet-surge strata deposited onto Early
199 Miocene sandstones of the Fil'akovo Formation. Superimposed is the succession with well-
200 stratified fine- to coarse-grained palagonitized lapilli tuffs with variable proportions of vesicular
201 basalt and occasional fragments of Miocene sediments. The alternation of finely laminated,
202 massive, unsorted and sorted coarse strata indicates dry-surge, wet-surge and pyroclastic fall
203 depositions. Both maars of Fil'akovo have been thought to be coeval with the Pleistocene
204 Günz/Mindel boundary (Konečný et al. 1999). However, U-Pb zircon combined with (U-

205 Th)/He zircon and apatite data ([Hurai et al. 2013](#)) returned an Upper Miocene-Lower Pliocene
206 age of the Castle Hill maar (5.62 ± 0.04 Ma, 5.5 ± 0.6 Ma and 5.4 ± 0.7 Ma, respectively). A
207 sample from the Red Hill maar investigated in this study has been excavated from the eastern
208 outer slope of the tuff ring of incoherent volcanoclastic material intercalated within siltstone
209 ([Figs. 2e, f](#)).

210 Heavy mineral assemblage of the volcanoclastic sediments under study also involves
211 pyroxene, amphibole, garnet, titanite, olivine, apatite, rutile, epidote, sillimanite and andalusite.
212 Garnet, amphibole and pyroxene dominate in the Fil'akovo–Red Hill maar, whereas sediments
213 from Tachty and Šurice diatremes are enriched in pyroxene and olivine ([Nemec and Huraiová](#)
214 [2018](#)).

215

216 **4. Methods**

217

218 Heavy mineral fractions hydro-dynamically separated from 10-20 kg of maar sediments
219 were further processed by sieving, gravitational separation in heavy liquids and final
220 electromagnetic purification. Zircons were hand-picked from diamagnetic fraction under
221 binocular microscope, mounted in epoxy resin, ground and polished to expose crystal interiors.

222 Zircon chemical compositions supplemented by cathodoluminescence and back-scattered
223 electron imaging were measured using a Cameca SX-100 electron microprobe at the State
224 Geological Institute in Bratislava (Slovakia), using 15 keV accelerating voltage, 20 nA beam
225 current, counting times between 5 and 25 s and 2-5 μm beam diameters. The instrument was
226 calibrated with natural and pure synthetic compounds as described in [Hurai et al. \(2012\)](#). Matrix
227 effects were resolved using the X-PHI correction method ([Merlet 1992](#)). Interferences were
228 eliminated using empirical correction factors ([Amlı and Griffin 1975](#); [Konečný et al. 2004](#)).
229 Detection limits were between 0.01 and 0.05 wt. %. Uncertainties varied from 1 % (major
230 elements) to 10 % (trace elements).

231 U-Th-Pb isotope data on zircons were obtained *in situ* by laser ablation inductively coupled
232 plasma spectrometry (LA-ICP-MS) at the Laboratoire Magmas & Volcans (LMV, Clermont-
233 Ferrand, France). The analyses involved the ablation of minerals with a Resonetics Resolution
234 M-50 Excimer laser system operating at a wavelength of 193 nm with a repetition rate of 3 Hz
235 and a laser fluency of 2.5 J/cm². The laser spot diameter was set at 27-40 μm ([Appendix S1.1](#)).
236 The ablated material was carried by helium and mixed with nitrogen ([Paquette et al. 2014](#)) and
237 argon before injection into the plasma source of a Thermo Element XR sector-field HR-ICP-
238 MS. The analytical protocol was similar to that reported by [Mullen et al. \(2018\)](#).

239 Data were corrected for U-Pb fractionation during laser sampling, and for instrumental
240 mass discrimination (mass bias) by standard bracketing with repeated measurements of GJ-1
241 zircon ([Jackson et al. 2004](#)). The 91500 zircon reference material ([Wiedenbeck et al. 1995](#)) was
242 also repeatedly analyzed as an unknown to independently monitor the reproducibility and
243 accuracy of the corrections. Data reduction was carried out with the GLITTER® software
244 package from Macquarie Research Ltd ([van Achterbergh et al. 2001](#)). For zircons younger than
245 10 Ma, ²⁰⁶Pb/²³⁸U ratios were corrected for the ²³⁰Th disequilibrium ([Schärer 1984](#)), assuming
246 a Th/U ratio for the whole rock corresponding to 3±1 ([Paquette and Mergoïl-Daniel 2009](#)). Data
247 were not corrected for common Pb. Calculated ratios were exported, and concordia ages and
248 diagrams were generated using the Isoplot/Ex v. 2.49 software package ([Ludwig 2001](#)).

249 Hafnium isotopes were measured with a Thermo Scientific Neptune Plus multi-collector
250 ICP-MS coupled to a Resonetics M50E 193 nm excimer laser system at the LMV, Clermont-
251 Ferrand, France. Data were collected in static mode (¹⁷¹Yb, ¹⁷³Yb, ¹⁷⁴Hf, ¹⁷⁵Lu, ¹⁷⁶Hf-Yb-Lu,
252 ¹⁷⁷Hf, ¹⁷⁸Hf, ¹⁷⁹Hf). The operating conditions and instrument settings are described in [Moyen](#)
253 [et al. \(2017\)](#), [Paquette et al. \(2017a\)](#) and reported in the [Appendix S1.2](#). Mass bias effects on
254 Hf were corrected using an exponential law and true value for ¹⁷⁹Hf/¹⁷⁷Hf = 0.7325 ([Patchett](#)
255 [and Tatsumoto 1980](#); [Patchett et al. 1981](#)). Data reduction was based upon user-selected
256 background and sample integration intervals. Outlier Hf isotopic ratios for each analysis were

257 rejected using a two-standard deviation criterion. The $\varepsilon_{\text{Hf}}(t)$ value is defined as the deviation
258 of the sample $^{176}\text{Hf}/^{177}\text{Hf}$ from the CHondritic Uniform Reservoir (CHUR) in parts per 10^4
259 calculated according to the $\lambda^{176}\text{Lu}$ value = $1.865 \times 10^{-11} \text{ yr}^{-1}$ (Scherer et al. 2001) and the CHUR
260 parameters ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282793$; $^{176}\text{Lu}/^{177}\text{Hf} = 0.0338$) from Iizuka et al. (2015).

261

262 **5. Results**

263 **5.1. Zircon ages**

264

265 Several zircon groups have been distinguished according to their U-Pb ages (Appendix
266 S2). Late Pliocene zircons represent the most abundant population (almost 30 % of all zircons)
267 in Tachty and Šurice diatremes, but missing in the Red Hill maar. The Pliocene zircons form
268 rounded and isometric grains, up to $1400 \times 1700 \mu\text{m}$ in size, which often represent fragments of
269 even larger crystals (Fig. 3). Corrosion was observed on the surface of smallest crystals. Dense
270 oscillatory or domain zoning of Pliocene zircons are discerned in cathodoluminescence (CL),
271 while inherited cores have not been observed. Crystal faces preserved in one grain (TA-2-1-2)
272 were consistent with a simple combination of (101) pyramid and (100) prism. The zircon
273 growth graded from the homogeneous central part to the rim crowded with innumerable thorite
274 inclusions. The texture, reminiscent of that observed in zircon cores from Nb, REE-rich calc-
275 alkalic granite xenoliths (Huraiová et al. 2017a), indicates the breakdown of $\text{ZrSiO}_4\text{-ThSiO}_4$
276 solid solution. The sharp boundary between the two zones rules out a post-growth, fluid-
277 induced dissolution-precipitation. Similar syn-growth textures have been observed in zircon
278 megacrysts from the Fil'akovo–Castle Hill maar (Hurai et al. 2013). The Late Pliocene zircons
279 yielded ^{230}Th -corrected U-Pb ages ranging from 3.0 to 2.2 Ma (Fig. 4). Zircons from Tachty
280 can be subdivided into three sub-groups clustering at $2.21 \pm 0.03 \text{ Ma}$ (24 analyses, 2 grains),
281 $2.37 \pm 0.02 \text{ Ma}$ (32 analyses, 3 grains), and $2.50 \pm 0.04 \text{ Ma}$ (12 analyses, 2 grains). Zircons

282 from Šurice cluster at 3.00 ± 0.07 Ma (5 analyses, 1 grain), 2.81 ± 0.06 Ma (6 analyses, 1 grain)
283 and 2.44 ± 0.02 Ma (37 analyses, 4 grains).

284 14 % of all investigated zircons returned Eocene-Miocene ages. One to three zircons of
285 this group are present in each locality. Compared to the previous group, Eocene-Miocene
286 zircons are smaller (typically 200-300 μm , occasionally up to 500 μm), almost euhedral
287 elongated crystals with fully or partially preserved, locally corroded faces. Typical dense
288 oscillatory and sector zonings are visible in CL. Some zircons contain rounded resorbed core
289 (CV-N2-1-5). The group of Eocene-Miocene zircons yielded ages clustering at 35.1 ± 0.6 Ma,
290 with additional individual data at about 38 and 29 Ma. One 35.1 Ma old zircon from Tachty
291 contains an inherited 306 ± 9 Ma core. Three zircon grains from Šurice and Fil'akovo-Red Hill
292 are concordant at 18.0 ± 0.3 Ma (Fig. 5a).

293 Palaeozoic zircons represent 25 % of the analyzed crystals. The Palaeozoic zircons are
294 more frequent in the Fil'akovo-Red Hill maar than in the Šurice diatreme. The group of
295 Palaeozoic zircons is texturally and morphologically rather heterogeneous, with smaller crystal
296 sizes, typically 200-250 μm . They form elongated crystals and fragments showing local
297 magmatic resorption and epiclastic corrosion. Dense oscillatory as well as patchy zoning of
298 cores grading into marginal oscillatory growth zones are discernible in CL images. All zircons
299 show discordant to concordant ages scattered between 300 and 510 Ma (Fig. 5b). A small group
300 of two discordant 300 Ma zircons from Fil'akovo-Red Hill is coeval with the inherited core of
301 the Eocene zircon from Tachty. Concordant ages span intervals from 504 ± 11 Ma (Early
302 Cambrian) to 373 ± 10 Ma (Late Devonian). In spite of the distinct zoning and clear core-rim
303 textural relationships, age contrasts have not been recorded within individual crystals.

304 Ages of Neoproterozoic zircons (23 % of investigated grains, Fig. 6a) span the interval
305 from 570 Ma (Ediacaran) to 850 Ma (Cryogenian). CL textures range from weak domain (700-
306 790 Ma) to dense oscillatory zonings (590-630 Ma). External morphology is isometric to
307 elongated, with signs of magmatic and chemical corruptions. An 854 Ma old zircon shows

308 preserved crystal faces. Most investigated grains are doubly terminated crystals. Some zircons
309 with sub-parallel parting and cataclasis have been excluded from isotopic study.

310 Rare Paleoproterozoic zircons (4 grains, 9 % of the investigated zircons) have been found
311 only in the Fil'akovo–Red Hill maar (Fig. 6b). Internal textures range from dense oscillatory
312 (e.g. FCV-Z1-3) to irregular domain zonings (CV-N2-3-1, CV-N2-1-6). Age of this zircon
313 group covers the interval between 1850 and 2115 Ma (Orosirian-Rhyacian). The oldest zircon
314 (FCV-Z1-5) exhibits a bright core-dark rim relationship expressed by two different ages, 2115
315 ± 18 Ma and 2065 ± 8 Ma, respectively. In spite of the complex internal zoning, ages of other
316 zircons were homogeneous and concordant. Multiple analyses of the FCV-Z1-3 zircon plot
317 along a discordia defining an upper intercept age of 1996 ± 14 Ma.

318

319 *5.2. Zircon compositions and Lu-Hf isotopes*

320

321 Pliocene zircons can be distinguished from older zircons in most discrimination diagrams
322 based on relative and absolute U, Th, Y, Yb, Zr, Hf abundances (Figs. 7, 8 and Appendix S4).
323 Zr/Hf and Th/U ratios are scattered within typical intervals of 40-90 and 0.5-3.5, respectively
324 (Appendices S2 and S4). High Th/U ratios, between 4 and 8, in zircons from Tachty (TA-8-1,
325 TA-1-1, TA-2-1-2, TA-Z1-2) are exceptional among all hitherto investigated SSVF zircons.
326 Zr/Hf and Th/U ratios in Eocene-Oligocene zircons are considerably lower, ranging at 32-33
327 and 0.1-0.6, respectively. Pre-Cenozoic zircons exhibit relatively low Zr/Hf ratios, ranging
328 from 30 to 50; whereas Th/U ratios are mostly significantly lower (0.1-0.8).

329 According to U-Pb and Lu-Hf isotope data summarized in Appendices S2 and S3,
330 respectively, zircons from basalt maars of SSVF create clusters corresponding to several
331 episodes of crustal growth (Fig. 9). Pliocene zircons differ from older zircons by unique super-
332 chondritic $\epsilon\text{Hf}(t)$ values, from 7 to 14, contrasting with sub-chondritic Eocene-Oligocene
333 zircons, with $\epsilon\text{Hf}(t)$ values ranging between -2.5 and -8. The pre-Tertiary zircon group forms

334 two discrete Palaeozoic-Neoproterozoic and Paleoproterozoic arrays, covering the interval of
335 $\epsilon\text{Hf}(t)$ values between -11.5 and +13.

336

337 **6. Discussion**

338 **6.1. Possible zircon sources**

339

340 Pyroclastic deposits of maar structures are composed of a mixture of volcanoclastic
341 material and disrupted rock fragments scavenged from feeder conduits. As such, they may
342 theoretically contain zircons from several sources: 1) oceanic crust and associated marine
343 sediments accreted to lithosphere during Miocene subduction (e.g. [Dobosi et al. 2003](#)), 2) pre-
344 Cenozoic basement rocks within Early Oligocene transgressive conglomerates, gravels and
345 sandstones ([Vass and Elečko 1992](#)), 3) Cenozoic sandstones of the Pannonian Basin ([Vass et](#)
346 [al. 2007](#)), 4) magmatic reservoirs occurring at various crustal levels (e.g. [Huraiová et al. 1996,](#)
347 [2017a](#)).

348 Detrital zircons involved in the Cenozoic sedimentary infilling of the Pannonian Basin may
349 also be provided by several distant sources: 1) uplifted Central Carpathian basement drained by
350 rivers, 2) ash-fall deposits from Eocene-Early Miocene volcanic centres. Areal Middle Miocene
351 andesite volcanism must have ejected a large amount of material distributed throughout the
352 whole Pannonian basin. In contrast, pyroclastic ejecta from small Late Miocene to Pleistocene
353 monogenetic volcanoes are probably negligible zircon suppliers owing to a smaller volume of
354 the volcanic material distributed over substantially shorter distances.

355 The proximal and distal zircon sources can be discriminated according to external
356 morphology of zircon crystals. Sharp crystal edges rule out longer sub-aquatic transport and
357 mechanical abrasion (epiclastic corrosion), thus excluding the Cenozoic sedimentary detritus
358 as the possible zircon source. Chemical corrosion of zircon crystals indicates the interaction
359 with hot volatile magma and the absence of sub-aquatic abrasion. Decompression in rapidly

360 ascending basalt may result in a specific parallel parting of zircon crystals (e.g. [Rudnick and](#)
361 [Williams 1987](#)). Rounded crystals indicate either magmatic resorption or the sub-aquatic
362 abrasion. However, zircons armoured in xenoliths may retain their external crystal morphology
363 until the ultimate phreato-magmatic explosion despite the longer transport in basaltic magma.

364

365 ***6.2. Origin of Cenozoic zircons***

366

367 Cenozoic zircons from SSVF maars cover the interval from Late Eocene (35-38 Ma) to
368 Pliocene (2.2 Ma). Their chemistry, internal zoning and morphology are unambiguously
369 magmatic in origin. Pliocene zircons from Tachty and Šurice diatremes create coherent age
370 clusters, which rule out their source in distal volcanic centres. The youngest zircon ages, i.e.,
371 2.2 Ma in Tachty and 2.4 Ma in Šurice, define lower limits for eruption ages. The Šurice
372 diatreme can thus be affiliated with the 3rd (2.45-2.75 Ma) volcanic phase, whereas the Tachty
373 diatreme belongs to the 4th (2.25-1.6 Ma) volcanic phase according to the scheme proposed by
374 [Konečný et al. \(1999\)](#) and [Konečný \(2007\)](#). The published 4.0 ± 0.3 Ma K/Ar whole-rock age
375 of basalt dyke intersecting the Šurice diatreme ([Konečný et al. 2003](#)) should be thus regarded
376 as erratic. The interval of 0.3-0.6 Ma between the oldest and youngest Pliocene zircons from
377 individual diatremes is consistent with previous studies ([Hurai et al. 2010, 2013](#)) and may
378 reflect the residence time in sub-volcanic reservoirs. The absence of post-Early Miocene zircons
379 younger than the surrounding basin deposits precluded the estimate of the eruption age of the
380 Fiřakovo–Red Hill maar.

381 Locally >1 mm megacrystic size, sector or oscillatory zonings and rounded shapes
382 attributed to magmatic resorption are typical diagnostic of the Pliocene zircons. Frequent
383 chemical corrosion reflects the interaction with volatile magma. Abundant crystal
384 fragmentation can be ascribed to the mechanical instability and destruction during phreato-
385 magmatic eruptions.

386 Zr/Hf ratios in the Pliocene zircons are scattered within the 40-90 interval, which is
387 consistent with megacrystic zircons from other maar structures (Hurai et al. 2013). The locally
388 increased Zr/Hf ratios (>60), significantly higher than those in zircons crystallizing from calc-
389 alkaline granites and pegmatites (Wang et al. 2010), may reflect a late Zr-saturation in more
390 alkaline parental melts (Boehnke et al. 2013). Th/U ratios in the Pliocene zircons range between
391 0.5 and 8.1, and overlap those of megacrystic zircons (0.2-4.6) from other maar structures
392 (Hurai et al. 2013). The high Th/U ratios coupled with increased Zr/Hf >60 are consistent with
393 an evolved alkaline syenite or phonolite parental melt (Belousova et al. 2002; Schmitt et al.
394 2010, 2017). The increased Zr/Hf and Th/U ratios in zircons from Tachty may reflect an
395 alkaline sub-volcanic reservoir (probably phonotephrite), which could fractionate to phonolite
396 in certain crustal levels at decreasing temperature.

397 Uranium abundances in Pliocene zircons mostly plot between the fields of zircons from
398 highly evolved, Nb-REE-rich A₁-type calc-alkaline granite xenoliths (Zr/Hf = 11-21, >1200
399 ppm U) and relatively less-evolved alkali syenite xenoliths (Zr/Hf = 38-74, 2-85 ppm U). The
400 decreased and strictly limited Hf contents (5,000-10,000 ppm), together with widely fluctuating
401 Y contents (380-11,680 ppm) are typical of felsic differentiation products from mantle-derived
402 mafic magmas (Pupin 2000). This zircon origin is also confirmed by the super-chondritic $\epsilon\text{Hf}(t)$
403 values (+7 to +14), diagnostic of the magmas derived by a low-degree partial melting of
404 lithospheric mantle and only weakly contaminated by an older crustal material. The $\epsilon\text{Hf}(t)$
405 values in the Pliocene zircons are in the lower range of ~5.5 Ma zircons (Fig. 9, inset) recovered
406 from A₁-type granites (Huraiová et al. 2017a). The close genetic relationship of megacrystic
407 zircons and alkaline syenites is also supported by the occurrence of yttrialite inclusions in
408 zircons from maar sediments and syenite xenoliths (Hurai et al. 2010, 2013, and unpublished
409 data).

410 The Late Eocene-Early Miocene zircons with U-Pb ages scattered between 38 and 18 Ma
411 form only a subordinate, but geologically interesting population of 1-3 grains in each locality

412 studied. Doubly terminated zircons show central domain encompassed by the zone with dense
413 oscillatory zoning. Dominant are (110) prisms combined with variable proportions of (211) and
414 (101) pyramids. The growth of (211) pyramid at the expense of the (101) pyramid was observed
415 in some crystals (e.g., TA-Z1-1, 38 Ma). The opposite trend was observed in 18 Ma old zircons
416 (e.g., CV-N2-1-5). The inherited 306 ± 9 Ma core in the Late Eocene, 35 Ma old zircon from
417 Tachty (TA-Z1-1) indicates recycled Carboniferous crust.

418 Dominantly euhedral shapes without signs of magmatic resorption and only moderate
419 chemical corrosion rule out the interaction of zircons with hot basalt as well as their mechanical
420 abrasive treatment in fluvial environment. Hence, the zircons must have been either armoured
421 in xenoliths during transport in magma, or scavenged from the feeder conduit of the maar
422 structure in uppermost crustal levels. Considering their scarcity, inconsistent ages, lacking
423 epiclastic as well as crystal-magma interaction features, the incoherent pyroclastic fall deposits
424 subjacent to the Miocene basin infilling are the most probable sources for these zircons.

425 Zr/Hf values in Eocene-Miocene zircons between 31 and 33 are consistent with a late-
426 magmatic origin (Wang et al. 2010). The dominantly anatectic parental magmas are also
427 corroborated by sub-chondritic $\varepsilon\text{Hf}(t)$ values between -2.5 and -8. The chemical and radiogenic
428 isotope fingerprints thus correspond to an orogenic, acidic-to-intermediate, subduction-related
429 magmatism. The 35-38 Ma old zircons are coeval with the volcanic centers along the
430 Periadriatic fault system and the middle-Hungarian zone interpreted as a subduction-related arc
431 volcanism (Kovács et al. 2007; Danišik et al. 2015). The obtained ages overlap the oldest
432 andesites from Velence and Récsk Mts. dated at 37 Ma (Benedek 2002).

433 The Early Miocene cluster at 18 Ma corresponds to the earliest syn-extensional eruptions
434 within the Pannonian Basin marked by silicic ignimbrite deposits of the Bükkalja volcanic field
435 south of Bükk Mountains. The ignimbrites are thought to reflect initial stages of the extension
436 accompanied by an increased heat flow from upwelled asthenosphere. Sub-chondritic $\varepsilon\text{Hf}(t)$
437 values in Bükkalja tuff zircons overlap those from SSVF maars and indicate the prevalence of

438 the crustal component in the parental granitic magmas, thus indicating initially thick continental
439 crust progressively thinned during the subsequent stages of silicic magma generation between
440 18 and 14 Ma (Lukács et al. 2015, 2018).

441

442 ***6.3. Origin and provenance of Palaeozoic-Proterozoic zircons***

443

444 Dense oscillatory and sector zoning similar to that in the magmatic Eocene-Oligocene
445 zircons is also typical for the Cambrian-Silurian zircons. The dominant (110) prisms associated
446 with (110) and (211) pyramids (e.g., SE-SP-1-1, SE-VR-2-4, CV-N2-2-1), together with Th/U
447 ratios between 0.2 and 0.6, and $\epsilon\text{Hf}(t)$ values between -2.2 and +2.2 favour the magmatic origin
448 and hybrid felsic parental melts.

449 Nearly all Ediacaran zircons exhibit the dense oscillatory zoning associated with high Th/U
450 ratios (up to 1.8). However, combinations of strongly negative and positive $\epsilon\text{Hf}(t)$ values (-8.4
451 in CV-N2-2-2 *versus* +12.8 in CV-N2-3-4) indicate synchronous occurrence of recycled crustal
452 protolith and juvenile parental melt.

453 Cryogenian zircons exhibit diverse zonings. Nevertheless, their high Th/U ratios (0.2-1.1)
454 combined with the super-chondritic $\epsilon\text{Hf}(t)$ values between +6.3 and +9.7 attest their close
455 genetic relationship with juvenile magmas.

456 Th/U ratios of four Early Proterozoic zircons range between 0.4 and 0.6, thus attesting their
457 magmatic origin. A typical “metamorphic” Th/U ratio of 0.01 (Rubatto 2002; Paquette et al.
458 2017b) was observed only in the 2115 ± 18 Ma core of the FCV-Z1-5 zircon, associated with a
459 super-chondritic $\epsilon\text{Hf}(t)$ value projecting close to the depleted mantle array. This probably
460 indicates the juvenile magmatic origin of the zircon grain overprinted by a solid-state
461 metamorphic recrystallization. Other Paleoproterozoic zircons project below the CHUR, thus
462 indicating the recycling of up to 3.4 Ga old mafic crust.

463 Assuming an evolution line of $^{176}\text{Lu}/^{177}\text{Hf} = 0.020$ characteristic of mafic crust (O'Neil et
464 al. 2013), most $\varepsilon\text{Hf}(t)$ values obtained from Oligocene to Neoproterozoic zircons can be
465 inferred from the continual recycling of ~1.0-2.0 Ga old crust mixed with a juvenile mantle-
466 derived component, probably during the Gondwana assembly (Fig. 9). The Paleoproterozoic
467 zircons indicate the Paleoproterozoic-to-Late Archean crust recycling during the assembly of
468 Columbia and its later incorporation in the peri-Gondwanan Cadomian belt.

469 Owing to large variations in trace element composition, Lu/Hf isotopes, and internal
470 zoning, the most probable sources of Palaeozoic to Proterozoic zircons must be sought in the
471 subjacent middle to upper crustal basement units (proximal source) and/or those cropping out
472 in Central Western Carpathians during the deposition of clastic fluvial sediments of the
473 Cenozoic basins (distal source). This assumption is also supported by the study of mineral
474 assemblages from the Red Hill maar, Tachty and Šurice diatremes, containing metamorphic
475 almandine garnets and rutiles derived from metasedimentary rather than metavolcanic
476 amphibolite-facies rocks (Nemec and Huraiová 2018).

477 U-Pb ages of pre-Cenozoic SSVF zircons overlap the Variscan, Caledonian, Cadomian and
478 Eburnean orogenic growth cycles recorded also in neighbouring tectonic units of Central
479 Western Carpathians. Rather surprising is, however, the very weak geochronological record of
480 large Upper Carboniferous granitoid plutons exposed in Central Western Carpathians, as well
481 as the lacking zircons coincidental with the acidic Permian-Triassic volcanism and the late-
482 Mesozoic and Early Paleogene metamorphic overprint of crystalline basements of the Alps and
483 Carpathians as well as the mafic granulite crust of the ALCAPA block. The gap between 850
484 Ma and 1.8 Ga indicates an interrupted crustal growth in the zircon source region, but this
485 behaviour may also be random, caused by the limited number of investigated zircons.

486 Palaeozoic zircons from SSVF with mixed super- and sub-chondritic $\varepsilon\text{Hf}(t)$ values indicate
487 a magmatic activity evolving from the Ordovician rifting to the Variscan collision. Relatively
488 abundant Neoproterozoic zircons indicate the evolution within the Cadomian orogenic arc

489 (Neubauer 2002; Linnemann et al. 2008), also documented by an inherited zircon core ($562 \pm$
490 43 Ma) in the Miocene zircon (13.2 ± 0.2 Ma) in the andesite laccolith near Fil'akovo (Bouloton
491 and Paquette 2014). The population of pre-Cenozoic SSVF zircons can be thus correlated with
492 an Armorican/Cadomian fragment of European Variscides (Putiš et al. 2008). Absence of
493 Mesoproterozoic ages might indicate the evolution of the source area of SSVF zircons within
494 the west-African Craton (e.g. Fernández-Suárez et al. 2004; Linneman et al. 2014; Henderson
495 et al. 2016).

496 Proximal and distal zircon sources cannot be unambiguously discriminated within
497 pyroclastic deposits. The absence of lithic fragments does not favour a proximal zircon
498 provenance in the subjacent crust sampled by the ascending magma along feeding channels.
499 However, alkalic syenite xenoliths representing the ultimate source of Pliocene zircons are
500 missing as well, in spite of the maximum abundance of these zircons in all investigated maar-
501 diatreme structures. Late Eocene-Early Miocene zircons derived from ash-fall deposits within
502 the basin infilling also favour the source of the pre-Cenozoic zircons in clastic sediments,
503 because blocks of the Miocene basin sediments are relatively abundant in the maar deposits.
504 On the other hand, sharp crystal faces of most pre-Cenozoic zircons rule out longer sub-aquatic
505 transport and mechanical abrasion, thus favouring a proximal (subjacent crust) rather than a
506 distal (uplifted Central Carpathian basement) provenance. The proximal source in the subjacent
507 crust should also be admitted for some Precambrian zircons showing the parallel parting
508 deformation textures reminiscent of zircons from rapidly decompressed crustal xenoliths
509 transported by ascending basalts (Rudnick and Williams 1987). However, the lower crustal
510 granulite protolith of the ALCAPA mega-unit can definitely be ruled out as the source of pre-
511 Cenozoic zircons, owing to the lacking high-temperature overprint and disturbance of the U-Pb
512 system. At any rate, the question of proximal or distal provenance of pre-Cenozoic zircon
513 xenocrysts in Pliocene maar structures remains open to further research.

514

515 **7. Conclusions**

516

517 Zircons recovered from pyroclastic deposits ejected by phreato-magmatic eruptions in
518 northern part of the Pannonian Basin are shown to provide unconventional information
519 pertaining to eruption ages of volcanoes and the nature of rocks scavenged from walls of
520 feeding channels of maar-diatreme structures. Most zircons have been derived from
521 intermediate to felsic differentiation products of intra-plate Pliocene alkali basalts. Owing to
522 minimum residence times (less than 300 ka) in sub-volcanic reservoirs, the Pliocene zircons
523 also define minimum eruption ages of the host maar-diatreme structures. Positive $\epsilon\text{Hf}(t)$ values
524 indicate a mantle-derived magmatic source uncontaminated by the crustal material. The Upper
525 Miocene-Pliocene magmatic zircons can safely be discriminated from older zircons by their
526 increased U/Th and Zr/Hf ratios.

527 Cenozoic zircons come from Late Eocene-Early Oligocene volcanic centres along the
528 middle-Hungarian zone, and the Miocene areal- and linear-type felsic-to-intermediate
529 volcanism. These zircons have been scavenged by the ascending basaltic magma from ash-fall
530 deposits from the base of the sedimentary basin infilling. Contrary to mantle-derived Pliocene
531 zircon antecrysts, their negative $\epsilon\text{Hf}(t)$ values indicate extensive crustal contamination or
532 recycling in the parental acidic-to-intermediate magma.

533 Pre-Cenozoic zircons returned ages scattered within the interval from Late Carboniferous
534 (~300 Ma) to Paleoproterozoic (Orosirian-Rhyacian, 1.8-2.1 Ga). Their variable internal
535 textures, heterogeneous sub- and super-chondritic $\epsilon\text{Hf}(t)$ values and trace element signatures
536 are diagnostic of the multiple recycling of continental crustal source. It remains unclear, if the
537 pre-Cenozoic zircons have been scavenged by the uprising basalt from the basinal detritus or
538 from the subjacent basement. Their source region could be correlated with an
539 Armorican/Cadomian fragment of European Variscides derived from the peri-Gondwanan
540 Cadomian arc and the west-African Craton.

541

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545

546 **Supporting Information**

547

548 Additional Supporting Information may be found in the online version of this article:

549 **Appendix S1.1** Details of analytical conditions for U-Pb isotope analysis

550 **Appendix S1.2** Details of analytical conditions for Lu-Hf isotope analysis

551 **Appendix S2** Zircon U-Pb isotope data

552 **Appendix S3** Zircon Lu-Hf isotope data

553

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840

841

Figure captions

842

843 **Fig. 1. a)** Schematic sketch of the Carpathian arc and the Carpathian back-arc basin (modified
844 after Pécskay et al. 2006; Kováč 2000): 1 – European platform, 2 – foredeep, 3 – Tertiary
845 accretionary wedge, 4 – Klippen belt, 5 – back-arc basin, 6 – Variscan crystalline basement

846 with pre-Tertiary sedimentary cover and Mesozoic nappe units (Cretaceous accretionary
847 wedge), 7 – Neogene andesite and rhyolite, 8 – Upper Miocene-Quaternary alkali basalt
848 volcanism, 9 – major normal, strike-slip and thrust faults. Rectangle marks the south-Slovakian
849 Volcanic Field (SSVF). **b)** Enlarged view of the SSVF with volcanic phases derived from
850 whole-rock K/Ar data on basalts (Konečný et al. 2003; Konečný 2007), U/Pb zircon, and (U-
851 Th)/He apatite dating of maars and diatremes (Hurai et al. 2013).

852

853 **Fig. 2.** Photo-documentation and geological cross-sections (Konečný and Lexa 2007) of
854 sampling sites. **a-b)** The Šurice diatreme with subsided block of the maar infilling in the upper
855 part. Explanations: 1 – palagonitized tuff breccia with fragments of Early Miocene sedimentary
856 rocks, 2 – subsided block of a maar complex: a – ash tuff, b – fine tuff, c – coarse tuff, d –
857 palagonitized tuff breccia, e – lapilli tuff with scoria fragments, f – agglutinated bombs in scoria
858 matrix, 3 – dykes and short sills, 4 – sandstone of the Fil'akovo Formation, 5 – scree. **c-d)**
859 Subsided maar deposits within the Tachty diatreme. Explanations: 6 – deformed bedded fine
860 palagonite tuff, 7 – palagonite tuff: a – fine, b – coarse, 8 – palagonite tuff breccia, 9 – slumped
861 palagonite tuff blocks in sandy sedimentary matrix, 10 – breccia composed of blocks of
862 sedimentary rocks, 11 – Early Miocene sediments. **e-f)** Areal view (GoogleEarth, Google Inc.)
863 and photo-documentation of the sampling site on the eastern flank of the Fil'akovo – Red Hill
864 maar.

865

866 **Fig. 3.** Morphological and textural types of Pliocene zircons from Tachty (TA) and Šurice (SE).
867 Scale bars (1000 μm) pertain to all images. BSE and CL refer to back-scattered electron and
868 cathodoluminescence images, respectively.

869

870 **Fig. 4.** Tera-Wasserburg plot of Pliocene zircons from **a)** Tachty diatreme (TA), Tachty -
871 playground (TA-Z) and **b)** Šurice diatreme (SE). Error ellipses display 2σ uncertainties.

872

873 **Fig. 5. a)** Tera-Wasserburg plot of Eocene-Miocene zircons from Tachty (TA), Šurice (SE) and
874 Red Hill (CV) with CL images of analysed zircon grains. **b)** Concordia plot of Palaeozoic
875 zircons from Fil'akovo–Red Hill maar (CV, FCV) and Šurice diatreme (SE) with CL images of
876 selected grains. Scale bars correspond to 100 μm . Error ellipses display 2σ uncertainties.

877

878 **Fig. 6. a)** Concordia plot of Neoproterozoic zircons from Šurice (SE) and Fil'akovo – Red Hill
879 (CV, FCV) with CL images of selected grains. **b)** Concordia plot of Paleoproterozoic zircons
880 from Fil'akovo–Red Hill maar with CL images of analysed grains. Scale bars correspond to 100
881 μm . Error ellipses display 2σ uncertainties.

882

883 **Fig. 7.** U (ppm) versus Th/U plot from LA-ICP-MS analyses ($n = 116$) of **(a)** Tertiary and **(b)**
884 pre-Tertiary zircons complemented with data ($n = 553$) from Hodejov, Fil'akovo–Castle Hill,
885 Gemerské Dechtáre (Hurai et al. 2013) and Hajnáčka–Bone Gorge maars (Hurai et al. 2012).
886 Zircon analyses from syenite ($n = 85$) and granite ($n = 29$) xenoliths (Hurai et al. 2010; Huraiová
887 et al. 2017a) are also shown for comparison.

888

889 **Fig. 8.** Discrimination plots constructed from EPMA analyses of Tertiary **(a, c)** and pre-
890 Tertiary **(b, d)** zircons complemented with the analyses of Late Miocene-Pleistocene zircons
891 from Hodejov, Fil'akovo–Castle Hill, and Gemerské Dechtáre maars (Hurai et al. 2013), and
892 zircon data from syenite and granite xenoliths from Pinciná (Hurai et al. 2010) and Čamovce
893 (Huraiová et al. 2017a).

894

895 **Fig. 9.** Hf-isotope evolution diagrams for SSVF zircons. Dotted lines in **a)** show the Hf-isotope
896 growths of 1, 2.2 and 3.4 Ga old mafic crusts with the $^{176}\text{Lu}/^{177}\text{Hf} = 0.02$. Depleted mantle curve
897 was constructed using the present-day depleted mantle ϵHf value = 16.9 (Griffin et al. 2002).

898 The detail in **b)** shows data on Tertiary detrital zircons from maars and diatremes, and those
899 from the Late Miocene (5.5 Ma) A₁-type granite from Čamovce ([Huráiová et al. 2017a](#)).